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AN EXAMINATION OF THE EVOLUTION OF RADIATION AND ADVECTION FOGS

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A literature study was done on radiation and advection fog evolution. For radiation fog, six stages of fog evolution have been identified -- 1) precursor, 2) sunset, 3) conditioning, 4) mature, 5) sunrise, and 6) dissipation. The evolution of advection fog models has been in parallel with radiation fog models, but no identified stages in the evolution of advection fog have been proposed in the literature. For advective fogs four stages of evolution are proposed: 1) precursor, 2) initiation, 3) mature, and 4) dissipation. Radiation and advection fog models will require greater sophistication in order to study fog spatial and temporal variability. Physical aspects that require further study are discussed.

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1. INTRODUCTION

Radiation type fogs are generally classified into ground, high-inversion, advective-radiative, upslope, and mountain-valley (Jiusto, 1981; Cotton and Anthes, 1989). The physical mechanisms responsible for the formation of fog involve three primary processes--(1) cooling of air to its dew point, (2) addition of water vapor to the air, and (3) vertical mixing of moist air parcels having different temperatures. For radiation fogs, the radiative cooling of air to its dew point is one of the primary processes, although the other two processes can contribute to its development and maintenance. Radiation fogs are generally classified as warm type fogs because the temperatures are above freezing; but under winter conditions, the fogs may be mixed with supercooled conditions near the surface and warm air conditions above the ground (Jiusto and Lala, 1983a, b, c). For advection fogs, processes (1) and (2) are more dominate. These are generally warm fogs, except in the colder climates where the temperatures are below freezing and ice crystals form instead of water droplets, producing an ice fog.

The question of how radiation and advection fog forms and evolves has been investigated for the past several years. Radiation and advection fog field programs have supplied information on many aspects of the problem, and fog diagnostic and prediction models have developed in sophistication so that they can reproduce fairly accurate one- or two-dimensional simulations of the fog evolution. Yet, no model has adequately simulated all four-dimensional physical aspects of radiation and advection fogs. The apparent failure of models to simulate "reality" may be traced to shortcomings in measuring and understanding the boundary layer processes, the fog microphysics, the surface heat and moisture budgets, advection, the influence of terrain, turbulence, and the radiation fluxes.

The purpose of this report is to review past field and model studies of radiation and advection fog and to present a fairly comprehensive survey of the present understanding of fog formation, persistence, and dissipation. This survey will help identify strengths and weaknesses of current models so that the necessary improvements can be made to understand and ultimately predict radiation and advection fog behavior.

Radiation fog forms during high pressure and clear skies after nocturnal surface cooling lowers the temperature to the dew point and sufficient condensation nuclei are available in the lower atmosphere. During the fog development, radiational cooling of the initial fog droplets and surrounding air causes further development and thickening of the fog. However, the exact process of fog formation is still being critically investigated. The role of turbulence is still being disputed. Some observations (Roach et al., 1976) suggest that fog forms after windspeeds decrease to low values and turbulence is nearly absent; and other observations (Rodhe, 1962; Lala et al., 1982) suggest that turbulence in the early evening may inhibit fog formation, whereas later in the evening it can help intensify fog.

Soil heat and moisture flux, dew formation and evaporation, condensation nuclei, and fog droplet settling are other factors that field and model experiments have verified to be important in fog formation, persistence, and dissipation processes. Also of interest have been the observations of quasi-periodic

oscillations in long-wave radiation, surface temperature, windspeed, and other parameters during the life cycle of radiation fog (Welch, Ravichandran and Cox, 1986; Roach, 1976; Lala et al., 1978; Gerber, 1981; Duynkerke, 1991). These oscillations, which vary from 31 s to 1 h, have been explained by variations in terrain, gravity waves propagating near the fog top, the balance between radiation cooling and turbulent diffusion, and advection of fog cellular (Bernard) cells (Roach, 1976; Welch and Wielicki, 1986; Choularton et al., 1981; Duynkereke, 1991).

An important and difficult aspect of fog formation modeling is the inclusion of radiative transfer processes due to aerosols, gases, water vapor, and fog droplets. Treating the radiative transfer problem requires the inclusion of multiple scattering, absorption and emission by fog droplets, water vapor, and gases. An exact procedure requires evaluating the radiative transfer equation over numerous spectral bands for each time and height step of integration (Korb and Zdunkowski, 1970).

A few sophisticated models have been fairly successful in approximating the heating/cooling rates of the atmosphere by parameterization of the absorption/transmittance and scattering processes because of gases, water vapor, and fog droplets (Zdunkowski and Barr, 1972; Brown and Roach, 1976; Roach and Slingo, 1979; Musson-Genon, 1987; Bott, 1991). However, multiple scattering by fog droplets remains a formidable problem and solutions have assumed idealized fog droplets (Zdunkowski et al., 1982).

Advection fogs form when warm, moist air moves over a colder surface such as land, water, and ice/snow. The surface must be sufficiently cooler than the air above so that the transfer of heat from air to surface will cool the air to its dew point and produce fog. Radiation and the other processes common to radiation fog may contribute to the formation of advection fog, but this type of fog requires the air to move from one place to another. The windspeed can be several meters per second, and is not as critical a factor as it is during radiation fog formation.

2. FOG FIELD STUDIES

Radiation and advection fog field programs have investigated many of the physical aspects of fog evolution. Emphasis has been on gaining a better understanding of fog formation, the role of various physical mechanisms, the fog microstructure (fog droplets, supersaturation, condensation nuclei, etc.), and dissipation. Sections 2.1 and 2.2 briefly summarize field program activity during the past 30 yr.

2.1 Radiation Fog

During the past 30 yr, radiation fog field programs have collected a number of data sets on boundary layer meteorology, cloud (fog) microphysics, and optical and aerosol physics variables (table 1). Some of the many fundamental questions addressed, but not necessarily answered, by these field programs have concerned the role of turbulent mixing in radiation fog formation, the effects of local advection, terrain features conducive to fog formation, the variations in liquid water content (LWC), the changes in visual range (optical extinction), variations

in drop-size distributions, fog supersaturation, heat and moisture budgets during the nighttime period, influences of air pollution on fog formation, and the role of solar radiation and cloud cover in fog dissipation.

Table 2 shows the fundamental variables or factors that were measured or addressed during selected field programs. The variables are generally classified into three broad categories--fog microphysics, optical properties, and boundary-layer meteorology.

The objectives of the field programs have generally varied in purpose, but the majority have been conducted to provide a better understanding of the full life cycle of radiation fog and to provide databases for evaluating and verifying the results of radiation fog models.

Radiation field programs have been conducted worldwide, and one of the typical field studies, FOG-82, was conducted at Albany County Airport, Albany, New York, from 7 September to 7 November 1982. In this study, the instrumentation and measurements provided a more complete description of the boundary layer, fog microphysics, and aerosol parameters than previously obtained from field studies (Jiusto and Lala, 1983a, b, c; Meyer et al., 1986).

The 1982 fog season was representative of the local fog climatology, with 22 h of measurements made during heavy fog conditions. Some of the results of the study are briefly summarized below:

- Five distinct stages in radiation fog evolution were identified: sundown, conditioning, mature fog, sunrise, and dissipation.

- Turbulence and associated vertical mixing can prevent, promote, or intensify radiation fog, depending on the fog evolution stage.

- Local heat and moisture advection may also play an important role.

- Observations showed that some fogs seem to form aloft initially and mix downward (high LWC aloft at 10 m early in the fog development). However, in the "conditioning stage" patches of ground fog occur below 1.5 m.

- The drop sizes in inland radiation fogs are larger than previously measured. Drops over 35 μm diameter were not uncommon. Also, the LWC can approach or exceed 0.5 g/m^3.

More recent results from the FOG-82 experiments have been published by Fitzjarrald and Lala (1989), who identified two new phenomena in relation to fog development. These phenomena are the following: (1) a jump in specific humidity occurs during the early evening transition that shortens the time required to reach surface layer saturation; and (2) for valley situations, an along-valley wind maximum near 100 m altitude is shown to be frequent, and its occurrence is associated with a threshold value of the along-valley surface pressure gradient. Such valley "jets" may influence the possibility of deep fog, depending on the heat and moisture advection.

This report does not contain detailed summaries of other field programs, but their pertinent information will be incorporated into the discussion on the evolution of radiation fog presented in section 4.

2.2 Advection Fog

During the past 20 yr, 19 field studies on advection fog have been identified from the literature (table 3). The studies have generally focused on the east or west coast of the United States, with a few overseas studies. The field studies have examined several aspects of advection fog including visibility, extinction and attenuation in fog, cloud condensation nuclei (CCN) and fog droplet spectra, supersaturation, turbulence, microstructure, threshold conditions for fog formation, the conditioning stage, temperature inversions, droplet sedimentation, and fog dissipation.

The CALSPAN studies along the coast of California and Nova Scotia certainly provided a significant database for advection fog modeling and microphysics. With the exception of the few studies in Russia and the one study in Tennessee, there is a significant lack of field studies of advection fog over land surfaces. This may, in part, be reflected in a lower frequency of advection fogs over land compared to radiation fogs and advections fogs near coasts, and possibly in classifying radiative-advection fogs as radiation fogs instead of advective.

The important aspects of some of the field studies are incorporated into section 5 on the evolution of advection fogs.

3. FOG NUMERICAL MODELS

Radiation and advection fog models have been developed at least since 1963 and have varied considerably in sophistication, depending on the objectives and purpose of the model study. Models may have existed before 1963, but the greatest effort in model construction has been after the 1960's, which has been done in phase with the sophistication of computer development.

3.1 Radiation Fog Model Studies

Table 4 shows a review of the radiation fog models that have been developed since 1963. About 35 models have been specifically developed to study radiation fog evolution. Zdunkowski and associates (table 4) have significantly contributed to radiation fog modeling. These developments have progressed from 1963 to the present. Also, the England group of Brown, Roach, and Associates (1976 to 1987) and the CALSPAN group (1972 to 1975) have significantly contributed to radiation fog modeling. The radiative transfer schemes have varied from simple parameterization of the radiative heating/cooling term to more complex schemes, including parameterization of fog droplet absorption, emission, and multiple scattering over the relevant wave lengths.

Radiation fog models have increased in complexity. Table 5 shows a required list of variables or factors to consider in modeling and how the past models have or have not incorporated these factors. As shown in table 5, the majority of models has been one-dimensional, with only limited efforts toward developing a fully three-dimensional model. Smolarkiewicz and Fitzjarrald (1988) suggest that one-dimensional models are too limited to provide enough information on understanding radiative fog processes. However, much of the past model development has focused on specific problems in regard to accurately simulating various physical processes such as the turbulent diffusion, radiation fluxes, heat and moisture

budgets, fog droplet growth and settling, and the effects of aerosol size distributions. Results from these models have also been used to compare with various aspects of radiation fogs as obtained from field measurements.

A detailed summary of all the models is not possible and is not the purpose of this report, but, like the field studies, the results of the various models will be incorporated in the discussion of section 4.

3.2 Advection Fog Model Studies

The development of advection fog models has essentially paralleled the development of radiation fog models, but apparently with less activity. Only about 12 advection fog models have been identified from the literature, but this may not include all of the overseas developments. The problems have been similar, and the focus has been on how to simulate various important processes or conditions such as condensation nuclei, supersaturation, turbulent exchange coefficients, radiation flux divergence, fog droplet sedimentation, horizontal advection, and initial conditions.

A list of advection fog models is given in table 6. Most of these models are one dimensional, with a few two dimensional. Fog models have been categorized by Saxena and Fukuta (1982) into "thermodynamic" and "kinetic" models. The first category assumes a state of thermodynamic equilibrium between fog liquid water and water vapor. Such models do not allow development of supersaturation, and are useful for identifying the fog forming conditions such as estimating fog LWC. The second category of models deals with interactions between the thermodynamically generated supersaturation field and the related microphysical processes, and are capable of estimating microphysical conditions of fog droplets such as visibility.

The CALSPAN advection fog model appears to be among the first to be developed (table 6). This two-dimensional model was developed to investigate the formation of advection fogs and their dissipation by natural and artificial heating. The model was based on several features of an earlier one-dimensional radiation fog model (Pilie et al., 1972).

The next significant development in advection fog models came with the studies of Hung, Liaw, and Vaughan (1980). A numerical model that described the evolution of potential temperature, water vapor content, LWC, horizontal and vertical winds, radiation cooling, growth of fog droplets, and droplet sedimentation was developed to study the differences in the characteristics of the formation of advection fog between CCN associated with polluted and clean atmosphere.

A number of Russian, Chinese, and other overseas authors have published results on their model developments (Sun et al., 1991; Ohnogi and Shibata, 1986; Khvorost'yanov, 1982; Buykov et al., 1981). Language difficulties prevent a complete understanding of these developments, but those available mainly show an interest in examining the physical mechanisms involved in advective fog formation and dissipation.

As with the other previous sections, the pertinent results from the various model studies will be discussed in section 5.

4. THE EVOLUTION OF RADIATION FOG

The earlier field experiments by Pilié et al. (1975a, b) indicated that the evolution of radiation fog can generally be divided into three stages--formative, mature, and dissipation stages. Low (1975a, b) analyzed microphysical data from five fog cases to investigate the evolution of radiation fog. Low indicated that the formative stage can be difficult to determine, but it is not too difficult to separate formation from maturity and maturity from dissipation. The end of the formative stage and the beginning of maturity are usually marked by a maximum in droplet concentration, LWC, and supersaturation and by a minimum in visibility and nucleus concentration. The end of the mature stage and the beginning of dissipation are usually accompanied by a decrease in droplet concentration, LWC, and supersaturation and by an increase in visibility, a greater broadening of the drop-size spectrum, and a rapid increase in nucleus concentration. Low also indicated that the time rate of change of these parameters was important in distinguishing the three stages. The time rate of change can be sharp, moderate, or steady and depends upon the particular fog event and the prevailing meteorological conditions.

During the 1982 Fog project at Albany, New York, Jiusto and Lala (1983a, b, c) identified a fourth and fifth stage--the sundown and sunrise periods. They indicated that these fog stages are identified by dominant physical processes whose associated microphysical and/or thermodynamic properties are distinctly different from those preceding or following it. The five stages are now considered as 1) sundown, 2) conditioning, 3) mature, 4) sunrise, and 5) dissipation. Welch et al. (1986) studied models to determine how well a model could simulate the observed stages of fog development. The model results generally reproduced the five stages of fog evolution.

The results from FOG-82 also suggested that a sixth stage of the fog development may be apparent. This will be called the pre-cursor or initial conditions stage. Meyer and Lala (1990) showed that fog development depends on season and critical synoptic conditions. Fitzjarrald and Lala (1989) stressed the importance of initial conditions on the possibility of fog development in the evening and early morning.

A summary of the six stages using the combined results of field measurements and models leads to the following picture of the formation, persistence, and dissipation of radiation fog. The interplay of the many factors and processes is outlined in figure 1, as adapted from Mason (1982) but modified for this study.

4.1 Precursor Stage

4.1.1 Season

Radiation fog can form at any time or season as long as the conditions are favorable. Meyer and Lala (1990) show a distinct radiation fog season at Albany, New York, in the late summer/early autumn. Spatola (1972) shows a similar trend for valley radiation fogs in West Virginia. This is primarily due to a sufficient period of nocturnal cooling coupled with an adequate moisture supply.

4.1.2 Location

The location of the site may be more favorable for fog formation because of terrain, vegetation, and presence of aerosols. The CCN are usually sparse in maritime air (~ 100 cm^{-3}), more plentiful in continental air (~ 1,000 cm^{-3}), and abundant in urban air (> 5,000 cm^{-3}) (Twomey and Wojciechowski, 1969). The chemical nature of CCN also depends on location. In maritime areas, the CCN are mostly salt aerosols, while in industrialized areas energy-related fuel combustion generates large quantities of SO_x, NO_x, NH_4, and H_2SO_4. These productions are, in turn, transformed into aerosol particles through photochemical reaction (Hung and Liaw, 1981a, b).

City heat islands may not be favorable for radiation fog formation compared to rural areas. Berlyand and Zashikhin (1982) used a numerical model to evaluate the effects of a city environment on the development of radiation fog. They conclude that there are factors that promote and prevent radiation fog. In the city, the urban air pollutants provide larger numbers of hygroscopic CCN so that the water vapor can condense at relative humidities below 100 percent. However, the city heat island effect works in the opposite direction and is a significant factor in preventing the formation of fog.

4.1.3 Synoptic Types

Meyer and Lala (1990) indicated that as many as five critical surface synoptic patterns were responsible for initiating the radiation fog process at Albany, New York. The more typical synoptic scenario is high pressure, clear skies, sufficient radiative cooling, and adequate humidity. In this case, the five stages of fog development as described by Jiusto and Lala appear relevant. However, a second scenario may be just as favorable. This scenario consists of a front or storm moving through an area, precipitating, and then moving out of the area with rapidly clearing skies. Clearing skies accompanied by radiational cooling and decreasing windspeeds establish the conditions favorable for fog development. The sunset stage may or may not be relevant since it depends on the departure time of the storm.

Clearing skies are an important condition. The net nocturnal radiation from the ground is roughly proportional to the height of the clouds. High clouds (cirrus) do not affect the net loss of heat from the ground significantly, but low clouds below 3.0 km above ground level (AGL) will reduce the net heat loss by one-seventh or more and, in many cases, prevent radiational cooling so the surface temperature will not reach the dew point (Monahan, 1977).

4.1.4 Boundary Layer Conditions

Fitzjarrald and Lala (1989) stressed the importance of initial conditions. Their studies suggest the following conditions are important for fog development:

- The likelihood of saturation throughout the stable boundary layer depends strongly on events that occur on the afternoon preceding fog rather than on existing conditions at sunset or evolution of surface boundary conditions during the night.

- The initial relative humidity that determines fog onset time in the boundary layer is the result of afternoon convective mixing.

- The boundary layer must not experience appreciable dry advection over night.

- Adequate soil moisture should be present.

A study in England (Saunders, 1973) investigated the upper limit of the geostrophic windspeed on radiation fog formation. This limit was around 10 m/s, and within the range of 9 to 10 m/s there was an increasing probability for stratus to form above the surface instead of ground fog. Below 10 m/s, Saunders found that fog formation depends on location as well as wind direction.

4.2 Sunset Period

4.2.1 Radiative Cooling and Temperature

On a clear evening and within an approximate 2-h period centered on sunset, the emitting long-wave radiation (F_R and H_R) (see figure 1) cools the ground, which then cools the lower atmosphere by conduction. The radiative cooling occurs mainly through the 8- to 13-μm infrared atmospheric window. Radiative flux divergence in the moist air near the surface and the gaseous radiative cooling are the main contributors. Davis (1957) found that on clear, calm nights the layer of maximum water vapor concentration is between 50 and 150 cm above the surface. A temperature minimum is found near the 20-cm level. Unless modified by turbulent mixing, advection, or radiation, this layer of maximum water vapor concentration may be favorable for incipient fog formation.

The vegetative cover of the ground can affect the radiative cooling. Brown and Roach (1976) indicate that a grass surface will cool to a lower surface temperature than bare ground because of its small thermal capacity. Also, grass will partially shield the soil from radiative loss. Thus, the air over a grass-covered surface will radiate to a colder surface than that of bare soil, and greater cooling of the air will occur.

The surface temperature decreases rapidly during the first 2 h. The surface layer changes from a lapse condition just before sunset to a strong temperature inversion structure (~ 8 °C/16 m). The heat loss at the surface is mainly balanced by heat conducted upwards through the soil (F_s), as a result of the temperature gradient established within the first few centimeters. The turbulent flux of heat (F_e) towards the surface makes only a small contribution (~10%) to the surface heat budget because of the low level of turbulent mixing that develops with a stable temperature profile. As the surface cools, the warmer air within the lowest few meters radiates directly to the colder ground.

4.2.2 Wind and Turbulence

The development of the strong temperature inversion near the ground results in the development of a nocturnal boundary layer of just a few meters. Wind velocities and turbulence near the surface decrease, which can result in a low-level windspeed maximum (~5 to 10 m/s) at around 200 m. This wind maximum usually decreases with time as the inversion increases with height.

4.2.3 Humidity and Haze

Accompanying the formation of the temperature inversion, the surface layer relative humidities increase from around 50 to 95 percent in 2 or 3 h. Correspondingly, the visual range decreases to 20 or 30 km as hygroscopic aerosols absorb moisture and increase in size because of the relative humidity increase. Jiusto and Lala (1983a, b, c) report that around a visual range of 10 km, the LWC is about 3×10^{-3} g/m^3, and there is a steady increase in the number concentration of haze aerosols in the 1-μm size range but with no aerosols greater than 10 μm.

4.3 Conditioning Stage

This phase may last from 2 to 11 h, depending on the radiative cooling, wind and turbulence, relative humidity, and stability of the lower atmosphere. Initial fog development may take place either as ground fog, fog aloft, or both.

4.3.1 Fog Onset Time

According to Fitzjarrald and Lala (1989), the fog onset time (FOT) can be estimated by

$$FOT \sim \ln(RH_0)\left[(\epsilon L_v/R\bar{T}^2)\,(\partial T/\partial t)\right]^{-1}, \tag{1}$$

where RH_0 is the initial relative humidity, ϵ is 0.622, L_v is the latent heat of vaporization, R is the gas constant, $\bar{T}$ is the estimated mean temperature during the evening, and $\partial T/\partial t$ is the estimated local temperature change that can be expressed as

$$\partial T/\partial t = -\partial R_n/\partial t - \partial(\overline{wT'})/\partial z - v_h \nabla_h T - w(\partial T/\partial z - \gamma) + L_v(c-e), \tag{2}$$

where the terms are (1) the local change in temperature, (2) radiative flux divergence, (3) turbulent flux divergence, (4) horizontal advection, (5) drying or warming effects of large-scale subsidence, and (6) the net effect of condensation or evaporation.

When FOT is greater than the length of night (LON), then fog is unlikely, and when FOT is less than LON, then fog is likely. The assumptions involved with equation (1) are that the specific humidity be constant with time, that the dew deposition be neglected, and that there be no significant horizontal advection. The author found that using FOT to predict likely fog nights is only partially successful. However, of 34 nocturnal cooling cases, the FOT equation predicted surface layer saturation (FOT < LON) in all but one of the 27 "fog patches" and "fog" experiments. Over 60 percent of the predictions were within 2 h of the observed onset of fog patches. Less than half of the "no fog" cases were predicted (Meyer and Lala, 1990).

4.3.2 Radiative Cooling and Temperature

During this period, the air within the lowest few meters continues to cool by radiative cooling, producing radiative cooling rates up to 1 °C/h. The surface temperature inversion increases in height, resulting in a gradual increase in height of the boundary layer and a slight increase in turbulent mixing below the inversion. This mixing then leads to increasing cooling at higher levels as heat (F_e) is transported to the ground to be radiated to space (F_r).

4.3.3 Dew Formation

The region below the top of the inversion now begins to show rapid increases in relative humidity (> 97%), and further cooling can cause saturation leading to dew deposition and some drying of the lower air. Pilie et al. (1975a, b) indicate that dew deposition is responsible for formation of a low-level dew-point inversion that appears to be a necessary condition for initial fog formation aloft. The inversion may extend from 40 to 200 m above the surface. Lala et al. (1975a, b), Brown and Roach (1976), and Pickering and Jiusto (1978) view dew deposition as a "governor" on fog formation. For a given rate of radiative cooling, if the dew deposition rate and accompanying downward transport of moisture are large, fog formation may be inhibited. However, if the dew deposition at the surface is somewhat less, radiative cooling may be sufficient to initiate the formation of fog.

4.3.4 Winds and Turbulence

Whether or not fog forms depends upon the balance between radiative cooling, leading to saturation of the air, and turbulent diffusion of moisture to the surface that dries and warms the air. Apparently, turbulence may act either to promote or to prevent fog formation. Some evidence (Jiusto and Lala, 1980; Welch et al., 1986) suggests that turbulence (winds > 1-2 m/s) early in the evening may inhibit fog; whereas, later in the evening as the depth of cooling and humidification increases, light turbulent mixing can intensify fog. However, in the fog experiments and model studies in England, Roach et al. (1976) and Brown and Roach (1976) indicated that significant fog development occurred when windspeeds dropped below 0.5 to 1 m/s. They infer that as the windspeed decreases, turbulent transfer of moisture to the surface to form dew ceases. As a result, the moisture remains in the atmosphere, and as radiation cools the air, fog is formed. At higher windspeeds, vertical mixing of drier air may inhibit fog formation.

Fitzjarrald and Lala (1989) show results from FOG-82 experiments in the Hudson Valley that indicate that the turbulent moisture flux convergence leads to a jump in specific humidity and to an abrupt drop in temperature, each effect diminishing the initial saturation deficit to be overcome before fog can form. The period of turbulence-dominated cooling and moistening and the later period of radiatively dominated cooling are distinguished by a sign change in the mean temporal curvature (vertical profile) of the surface-layer temperature. This change in the curvature of the temperature trace and the jump in specific humidity is observed during the period when the turbulence has fallen to near zero.

In modeling, Lala et al. (1975a, b), Brown and Roach (1976), Zdunkowski and Barr (1972), and Welch et al. (1986) concur that the structure of fog and the occurrence and nonoccurrence of fog depend strongly upon the particular profile of eddy viscosity or turbulence used in models. However, Smolarkiewiez and Fitzjarrald (1988) suggest from their model studies that fog evolution is fundamentally a Benard convection problem, determined not so much by the surface layer turbulence but rather by the dynamics of the fog top interface, which in turn depends on the environmental structure (stability, shear, moisture) evolving continuously due to imposed volume sources/sinks of heat related to the infrared radiative transfer and water phase exchange.

4.3.5 Fog Formation and Droplets

Conditions under which fog formation occurs at the surface or aloft are still unresolved. Jiusto (1981) indicated that many inland radiation fogs first develop at the surface and then develop upward. However, Pilie et al. (1975a, b) indicate that fogs may develop aloft, especially in valley terrain. Model studies appear to show both types of fog formation.

Fitzjarrald and Lala (1989) distinguish two types of fog formation--surface-layer fogs and boundary-layer fogs. Surface-layer fogs are characterized by strong stability, a depth of about 20 m, and a variable visual range and usually precede boundary-layer fog. Boundary-layer fog usually extends to about 150 m, is strongly influenced by clear air radiative cooling and advection effects from local winds, and may modify winds and temperature in the boundary layer. Boundary layer fog may or may not be preceded by surface layer fogs.

When patches of fog occur, they are mainly surface phenomena (ground fog) and usually appear below 2 m. However, at times, fog can appear to form aloft (higher LWC aloft) and mix downward, or the fog may actually form in an elevated layer somewhere else and drift by advection (Ha) over the site.

At this stage, the visual range decreases (~10 km) because of light haze and occasionally decreases to 1 km or less because of light but nonpersistent fog. Visual range, relative humidity, drop concentrations and sizes, temperatures, winds, and computed Richardson's numbers (RI) are highly oscillatory. Lower visual range (fog) is associated with stable RI (> 0.5) and often with enhanced outgoing radiation (Jiusto and Lala, 1980). Lala et al. (1978) and Meyer et al. (1980) observed that visual range (or visibility) decreases with increasing large particle concentrations, and at a visual range of 1 to 2 km, their measurements indicated a transition from haze (< 1 μm) to larger fog droplets. Thereafter, the concentration of drops becomes somewhat less important than the size of fog drops in attenuating light.

Pilie et al. (1975a, b) show that ground fog is characterized by droplet concentrations of 100 to 200/cm^3 in the 1- to 10-μm radius range, with a mean radius of 2 to 4 μm. As deep fog forms aloft, droplet concentration near the surface decreases to less than 2/cm^3, and the mean radius increases from 6 to 12 μm. Droplets of radii less than 3 μm disappear. Thereafter, droplet concentration and LWC increase gradually until the first visibility minimum at the surface when typical values range from 12 to 25/cm^3 and 50 to 150 mg/m^3, respectively. The small droplets reappear at the first visibility minimum.

Supersaturation in the thin ground fog exceeds that in deep fog. The initial surface obscuration in deep fog appears to be due to droplets that form aloft and are mixed downward into unsaturated air by turbulent diffusion. New droplets are apparently not generated near the surface until after the first visibility minimum.

Usually in the later part of this stage, fog may form and the top will oscillate in height. The LWC may average around 0.05 g/m^3 near the fog top to about 0.15 g/m^3 near the ground. Radiative cooling rates are strongly correlated with the LWC in shallow ground fog.

During the conditioning state, at least three modes of aerosol or droplets may form. Jiusto and Lala (1983a, b, c) indicate that for visual ranges around 1 km and LWC of 0.03 g/m^3, the haze size mode around 1 μm becomes more pronounced and persists while a second size mode of 5 to 7 μm appears. A larger fog droplet size mode between 10 μm and 35 μm appears with supersaturated conditions.

The development of fog droplets has important radiative implications. Roach (1976) indicates that droplet growth can occur in a slightly subsaturated environment. This process allows for fog formation with consequent radiation-induced changes in the stability of the entire cloudy layer, even though the environment may never become supersaturated with respect to water on the average. Brown (1980) and Mason (1982) conclude that the importance of the radiation term in the droplet growth equation varies with the concentration of activated cloud CCN. At high CCN concentrations, the radiative term has little influence on either the mean droplet size or the LWC (more numerous droplets have small values of absorption efficiencies). They also concluded that the radiative exchange between droplets and their environment will be greatest in clean fogs and maritime layer clouds and less in heavily polluted air.

4.4 Mature Fog Stage

The period of "conditioning" of the vertical temperature and dew-point profiles and mixing of vertical parcels of moist air produce supersaturation and dense fog formation. Once dense fog (visual range < 0.4 km) is formed the time variations or oscillations in visual range become highly damped. Generally, other distinct changes are slightly stronger surface winds, very low RI, a lapse rate below the inversion that may vary from isothermal to superadiabatic, much higher droplet concentration, and a shift to larger drop sizes.

4.4.1 Supersaturation

The fog onset time derivation of Meyer and Lala (1990) and Fitzjarrald and Lala (1989) does not allow for supersaturation in fogs. There has been considerable discussion in the literature on fog supersaturation (Gerber, 1991). Maximum supersaturation in fogs is estimated by comparing the observed critical super-saturation spectrum of CCN in the fog forming air mass and the fog droplet number concentration (Fitzgerald, 1978; Hudson, 1980; Saxena and Fukuta, 1982). This "one-to-one correspondence method" estimates fog supersaturations on the order of or less than 0.1 percent. However, direct measurements of water vapor supersaturation using a saturation hygrometer (Gerber, 1982) and a droplet spectrometer (1982 Fog Project) showed values as large as 0.5 percent. The large

supersaturations may be one explanation for larger-than-expected droplets (~ 40 μm droplet diameter) in radiation fogs (Pinnick et al., 1978; Choularton et al., 1981).

Gerber (1991) uses the ideas of nongradient turbulent diffusion and the application of these ideas to the interaction of turbulence with fog droplets as a possible mechanism for explaining the larger supersaturations. The concept of nongradient mixing is used by Broadwell and Breidenthal (1982) as an alternative to the popular gradient diffusion method. They showed experimentally that mixing of two species in turbulent shear layers is a two-stage process where the identity of the species remains largely intact while being mixed throughout the turbulent layer by larger-scale inviscid motions, and where homogenization of the species occurred only by molecular diffusion near the interface created between the species during the mixing process. The rate of homogenization is minimal until the turbulence scale approaches the Kolmogorov microscale when the interface rapidly increases.

A model is developed by Gerber, and the results appear to support the nongradient mixing hypothesis. Gerber proposes that nongradient mixing of eddies causes supersaturation transients because the release of excess vapor by molecular diffusion at the interfaces of nearly saturated air mixing in eddies is faster than the relaxation time of droplet response to this excess. Thus, the nongradient mixing mechanism can be an effective means of activating new CCN entrained into fogs, as well as CCN already located in the interior of fogs. The effectiveness of this mixing depends on the value of the integral droplet radius and on the frequency and magnitude of temperature gradients in the near-saturated mixing environment.

4.4.2 Radiative Cooling and Temperature

As the fog increases in LWC and depth, radiative cooling of the fog droplets (H_r) becomes dominant in the upper part of the fog while the radiational loss from the surface (F_r), screened by the fog, is reduced to a value below that of the soil heat flux (F_s). Radiational cooling rates may reach -3 °C/h to -4 °C/h in the upper region of the fog.

Once the fog reaches a height of about 20 m, the radiational loss from the surface (F_r) decreases by about a factor of 2 and the surface temperature no longer decreases. As the fog grows thicker, the surface radiation loss (F_r) continues to decrease and the surface temperature begins to increase by about 1 °C/h. The lower atmospheric stability first becomes isothermal and then toward unstable conditions as the RI steadily decreases and windspeeds and turbulent mixing increase. The temperature inversion lifts from the surface and continued mixing can establish a moist or superadiabatic lapse rate near the ground and the onset of weak convective motion. Eventually, the inversion and the region of maximum wind shear migrate from near the surface towards the fog top, where their opposing influences determine the intensity of turbulent mixing across the fog top, and hence its sharpness. The smaller droplet sizes and the observed constancy of LWC with height in the upper part of the fog are evidence for strong mixing across the upper boundary (Mason, 1982).

The structure of a mature radiation fog was investigated by Brown (1987) at Cardington, England, during 17-18 October 1977. The structure was summarized by dividing the temperature profile into three regions. The first region is the portion of the inversion above the top of the dense fog. Radiative cooling to the cold fog top can lead to condensation and hence the upward growth of the fog. Turbulent mixing may also play an important role in the upward growth because of the elevated maximum in wind shear and the presence of gravity waves. The second region is the portion of the inversion containing the dense fog. The radiative cooling rate is around 3 °C/h, yet locally there is little temperature change. Factors that can offset the cooling include the mixing in of warmer air from above the fog, the release of latent heat, and the weak convective transport of heat from below. The third region is the well-mixed region below the base of the temperature inversion and within the fog layer caused by the weak convection generated by warming at the surface and radiative cooling at the top. Significant radiative cooling often extends beneath the second region.

4.4.3 Fog Droplets

In dense fogs, the fog drop size is rather broad, extending from a radius of 2 μm to about 30 μm. Some dense fogs appear to have bimodal distributions. Pilie et al. (1975a, b) report a drop-size distribution mode at 2 to 3 μm radius and one at 6 to 12 μm. Meyer, Jiusto, and Lala (1980) observed a primary mode at 5 to 7 μm radius, a secondary mode at 2 μm, and a weaker third mode at around 10 μm. Other studies (Mack and Pilie, 1973; Jiusto and Lala, 1980; Mason, 1982; Choularton et al., 1981) have reported relatively frequent drop sizes between 5 and 12 μm. Jiusto and Lala (1983a, b, c) indicate that in dense fog with visual range about 250 m and LWC of around 0.15 g/m^3, there is a sustained fog supersaturation and drops grow to larger sizes by diffusion, while there is an actual reduction in drop sizes around 2 to 5 μm radius. In very dense fog where the visual range is 75 m and LWC is around 0.5 g/m^3, the large drop mode increases in radius to 12 to 15 μm. Significant concentrations of drops can be found in the 20-μm radius range.

In the case of larger fog droplets, Roach and Slingo (1979) indicated that roughly half of the infrared radiation intercepted by the fog droplets is scattered, predominantly in the forward direction. However, in most of the spectra, the effect is small because the absorption by atmospheric gases is so large, but this situation is reversed in the 8- to 12-μm window region. Multiple scattering increases the effective absorber amount because radiation that would otherwise leave the scattering volume can be scattered back into it, and hence additional absorption can occur. The scattering by droplets increases the optical path length in the fog, resulting in additional absorption, and reduces the radiative cooling.

Roach (1976) examined the effect of radiative cooling on droplet growth and demonstrated the importance of droplet settling on controlling the water budget. He concluded that the maximum radius to which droplets would grow in a fog is around 10 μm.

These results are consistent with most observations, except for Pinnick et al. (1978) and Choularton et al. (1981) who found fog drops with radii > 15 to 25 μm.

Choularton et al. explain the existence of these larger fog drops on the basis of two possible mechanisms. Both mechanisms use the unstable and convective region beneath the fog top and inversion layer. The first mechanism requires the drops to grow because of large supersaturation fluctuations near the fog top, resulting from downwards entrainment of warmer, moister air into the colder air of the fog. The second mechanism uses the convection motions to cause a fraction of the larger drops to make several excursions to the radiative cooling region near the fog top.

Measurements of droplet concentrations indicate that they can vary as much as $12/cm^3$ to $> 300/cm^3$, and LWC can vary as much as 50 to 400 mg/m^3. The cloud nucleus concentration active at supersaturation of 3 percent is usually between 800 and $1000/cm^3$ near the surface, and decreases with increasing altitude and increasing age of the fog (Pilie et al., 1975a, b).

Model studies have shown the importance of fog droplet settling. If droplet settling is not included in models, unrealistically high LWC is predicted. Brown and Roach (1976) indicate that a feedback process is involved with fog droplet settling. The direct removal of liquid water by droplet settling causes a reduction of radiation cooling due to cloud droplets. The lower radiation cooling leads to a reduction in the rate of condensation and LWC. Brown (1980) suggests that fog LWC is sensitive to CCN concentrations. Lower concentrations of CCN were associated with fog LWC that was about 20 percent less than with higher CCN concentrations. The difference was attributed to reduced droplet settling in the high CCN situation.

Although dew formation appears rather important during fog formation, Pilie et al. (1975a, b) indicate that from the time of fog formation until sunrise, dew does not appear to serve any major function other than to maintain a saturated lower boundary for the fog.

4.4.4 Fog Deposition

Deposition during radiation fog episodes has been studied in the San Joaquin Valley of California, the Po Valley in Italy, and in Albany, New York (Waldman et al., 1987; Fuzzi et al., 1984, 1986). A numerical model for studying acidic deposition in radiation fog has been developed by Pandis and Seinfeld (1989).

Wet deposition, through sedimentation and impaction, can make a significant contribution to the overall flux of pollutants. When atmospheric stagnation prevents normal ventilation in a region, the buildup of atmospheric constituents will be governed by (1) primary emissions, (2) in situ transformation, (3) intrabasin circulation, (4) ventilation, and (5) removal by deposition to ground and other surfaces. The mixing height controls the volume in which these processes occur. During dense fog, deposition becomes the predominant loss term for secondary aerosol species.

Studies at Albany, New York (Fuzzi et al., 1984), show that the variation of LWC seems to be the leading mechanism for the chemical concentration changes in fogs with low to moderate levels of pollution. Fog water pH values at Albany ranged from 4.3 to 6.4 in contrast to reported values of 2.2 in Los Angeles, California, and 2.8 in the Po Valley of Italy.

In a study of radiation fogs in the southern San Joaquin Valley of California (Waldman et al., 1987), deposition rates for major chemical species were 5 to 20 times greater during fogs compared to nonfoggy periods. Sulfate-ion deposition velocities measured during fog were 0.5 to 2 cm/s. Rates measured for nitrate ions were generally 50 percent below those for sulfate, except for acidic fog (pH < 5) conditions. Scavenging of ambient aerosol was observed to increase as LWC increased. The lifetimes for atmospheric sulfate and ammonium were 6 to 12 h during dense fog compared to the ventilation rate of about 3 days for valley air.

A Lagrangian model was developed by Pandis and Seinfeld (1989) to study acidic deposition due to radiation fog. Results indicated that deposition rates of the major ions increased significantly during the fog episode, the most notable being the increase of sulfate deposition. Expressing the mean droplet settling velocity as a function of LWC is found to be quite influential in the model's prediction.

4.4.5 Quasi-Periodic Oscillations

During the mature fog stage, a series of fog dissipation and redevelopment episodes can occur (Welch and Wielicki, 1986). LWC increases in the upper regions of the fog during the quiet periods, and subsequent radiative cooling eventually causes destabilization of the lower atmosphere, increasing turbulence generation and mixing of the upper-level liquid water (F_d, J_d, F_v, J_v) to the surface, creating surface fog intensification. However, fog droplet settling (J_g, F_d, F_d), droplet impaction on vegetation, and dew deposition tend to decrease the fog liquid water near the surface.

Quasi-periodic oscillations in fog variables such as net radiation, surface temperature, wind, LWC, and visibility have been reported by Tahnk (1975), Roach (1976), Lala et al. (1978, 1982), Choularton et al. (1981), and Duynkerke (1991). The oscillations in net radiation, surface temperature, and wind were observed to be around 10 to 15 min, with the net radiation out of phase with the other variables. Lala et al. (1978) provide an interpretation of the mechanisms producing the phase relation of the variables as follows: an increase in windspeed augments vertical mixing that causes the surface temperature to increase and the relative humidity to decrease. The surface temperature lags because of the large heat capacity of the soil. The size of fog droplets decreases because of the reduction in relative humidity and evaporation. The net radiation decreases, resulting in the net radiation being out of phase with the other variables. Note that the passage of a layer of low clouds such as stratus can upset the phase relationship. In this case, the net radiation is in phase with all the variables, except for the windspeed, which is out of phase. After passage of the cloud layer, the earlier oscillations become evident again.

In the model experiments of Welch and Wielicki (1986), a series of pronounced oscillations in fog parameters was reported that have not been observed entirely in field observations. These oscillations occur about every 30 to 40 min and are the result of surges in turbulent eddy mixing. A surge in eddy mixing causes the height of the boundary layer to increase and later contract. When the boundary layer height increases, heat from the upper portion of the fog near the inversion top is mixed into the lower regions of the fog. Air temperatures in the lower 20 m increase at a faster rate than the surface temperature, resulting in a small surface inversion to form and suppress turbulent mixing.

The thick fog aloft radiatively shields the surface, preventing surface cooling. The surface warms, while radiative cooling of the fog causes a gradual decrease in air temperature. The net effect is the rapid breakdown of the surface inversion and the regeneration of turbulent mixing. Coupled with droplet settling, this increased turbulence causes the downward transport of liquid water from aloft. With another surge in turbulent mixing, the cycle repeats. These events act as a pumping mechanism to progressively transport moisture higher into the boundary layer. The fog top growth is now strongly correlated with surges in turbulent mixing. The fog continues to grow in height, and dry air aloft is mixed down and gradually decreases the maximum LWC found near the ground. Droplet settling due to size of drops and the turbulent mixing, also contributes to a decrease in fog liquid water.

Other explanations have been given for the observed oscillations. Roach (1976) and Duynkerke (1991) explain the quasi-periodic oscillations in terms of gravity waves propagating at the top of the boundary layer (fog top) or to a fluctuating balance between radiation cooling and turbulent diffusion. Model experiments by Smolarkiewicz and Fitzjarrald (1988) suggest that fog evolution is a Benard convection problem where the instability at the fog top interface and not the surface layer turbulence produces internal fog convection circulations. Welch and Wielicki (1986), using satellite observations, confirm that fog is composed of cellular elements, and the advection of these cells across a site may contribute to the observed oscillations. Choularton et al. (1981) observed periodic oscillations in LWC in a fog layer in West Germany, varying from 31 s to 96 s during 15 to 20 min intervals. They found that the period of gravity waves was around 250 s, much too long to account for the smaller oscillations. Measurements of the stability in the fog layer, the windspeeds (1.5 m/s), the depth of the fog, and Rayleigh number criteria suggested to Choularton et al. that the oscillations were due to convection of the Benard cell type being advected over the site. They assume that the cells have a broad upward airflow in the middle and narrow, stronger downward flow around the edges of the cell. Regions of high liquid content are correlated with the downdrafts of the cells.

4.5 Sunrise Period

At sunrise an additional variable enters into the fog equation, which is the solar short-wave radiation. In many cases, the heating of the surface will be sufficient to modify an existing radiation fog, but in some cases the fog may be so well established that the solar radiation is not sufficient to make much of an impression on the fogs persistence. In other cases, the solar radiation may actually help in the persistence or formation of fog.

4.5.1 Water Budget

After sunrise, the water budget in a radiation fog is altered considerably. During the night, water is deposited out of the atmosphere through the processes of dew formation, precipitation, and impaction. This liquid water acts as a moisture reservoir for evaporation into the fog after sunrise. The evaporation of liquid water from the surface of plants and objects back to the atmosphere dominates and can prolong the lifetime of the fog by helping to maintain saturation of the air as it warms.

Wattle et al. (1984) found that the maximum water deposition rates peaked between 0545 and 0645 LST coincident with the maximum in LWC. Dew collection rates in absence of fog averaged 18 $g/m^2/h$. In fog, water deposition rates averaged from 34 to 74 $g/m^2/h$, depending on height of vegetation. They suggest that dissipation of fog would have occurred somewhat sooner without the moisture source at the surface provided by the accumulated effects of dew formation, fog precipitation, and vegetative interception of fog droplets. Frost and freezing of collected dew did not measurably change the rate of vapor collection before the fog.

4.5.2 Vegetation

Individual vegetation elements serve as impaction sites for collection of significant quantities of fog droplets during dense fog, accounting for ~ 50 percent of the deposited water. For vegetation that stands substantially above the surface, fog water deposition rates are expected to be substantially greater (Wattle et al. 1984).

4.5.3 Fog Formation

Field measurements (Jiusto and Lala, 1983a, b, c) have indicated that dense fog can form just at sunrise. It is suggested that the late fog formation may be due to convection, enhanced mixing, and evaporation of dew. A late forming fog may be patchy and of relatively short duration (1 to 3 h). In addition, a fog already in existence can become more intense shortly after sunrise.

4.6 Dissipation Stage

Radiation fogs usually dissipate from 2 to 4 h during calm conditions. In some situations, fog can persist all day or regenerate again at night. The mode of dissipation can either be from the ground and lifting to form a low stratus cloud or from the top downwards. Most field observations and model studies have verified the lifting mode. Fog can be dissipated by 1) solar radiation heating the ground and convective warming from below, 2) direct absorption of solar radiation by fog drops, 3) evaporation or sedimentation of the larger fog droplets, 4) the advection of a low-level cloud layer over the fog, 5) increase in windspeed and turbulence, and 6) eroding of the fog edges by convective motions.

An indication of dissipation may be detected during the last hour of the fog's life cycle when a dramatic decrease in droplet concentration, LWC, and supersaturation can occur also, with an appreciable improvement in visibility, a greater broadening of the drop-size spectrum, and a rapid increase in nucleus concentration (Low, 1975a, b).

4.6.1 Solar Radiation

Solar heating causes the saturation vapor pressure to increase above the actual vapor pressure in spite of the evaporation of dew. However, the long-wave radiative cooling is not diminished within the fog as a consequence of the sunrise. Thus, for a while long-wave radiative cooling prevails over solar heating and the fog LWC continues to rise. Later, the solar heat flux prevails, causing the evaporation of fog liquid water, a reduction in fog density, and

eventual dissipation of the fog. Absorption of solar radiation by the fog droplets and the multiple scattering of the solar radiation can be contributing factors (Vehil and Bonnel, 1988). Another important factor is the gradient in temperature between the soil and the air. If the soil-air temperature gradient is small, fog dissipation may be unlikely. If it is large, fog dissipation is more likely, especially when a cloud layer moves over the fog layer.

4.6.2 Advection of Clouds

The advection of a cloud layer over a fog can alter the net divergence of long-wave radiation from the top of the fog, leading to a change in the fog structure and eventual dissipation of the fog. Saunders (1957, 1960a, b) observed that a large fraction of fog cases cleared following the arrival of a cloud layer. He also observed that the net upward radiation flux at fog top decreased substantially compared to clear sky, depending on the height and temperature of the overlying cloud layer. The lower the cloud layer height, the greater the reduction in net upward radiation flux. The reduction in the net upward radiative flux from the fog top along with the upward heat flux from the ground causes the air temperature to rise, resulting in the evaporation of the fog droplets.

Observations and model studies (Mason 1982) indicate that the advection of an overlying low cloud layer over the fog is more effective in dissipating fog than solar radiation. A cloud layer can clear a fog layer in an hour or two by reducing the radiative heat loss from the fog top (decreases the upward radiation from the fog top) and allowing the heat flux emanating from the ground and the low-level turbulence to disperse the fog. Although solar radiation is less effective in dissipation, because long-wave radiative cooling can continue from the fog top, the convective warming and heating of the ground eventually dissipate the fog layer by evaporation of the fog liquid water.

In many cases, the fog lifts from the surface and forms a low stratus cloud, but it has also been reported that fog layers have cleared from the top downwards. The mode of clearance may depend on windspeed and relative humidity of the air above the fog.

A model study by Welch et al. (1986) shows one possible mode of fog dissipation. The process starts with the moisture aloft decoupling from the surface and a low-level stratus cloud forming around 100 to 200 m. The stratus cloud radiatively shields the top of the fog layer below, damping oscillations in the fog. The fog rapidly dissipates to a ground fog while the stratus layer increases in height. The ground fog grows in height temporarily as dew and soil moisture is evaporated. The surface fog finally dissipates, but with lingering high relative humidity and haze. Meyer et al. (1980) report that during the fog dissipation stage, a fog droplet peak at 10 μm diameter persists, even after a stable haze condition is attained.

4.6.3 Convective Dissipation

Layers of fog that cover an area of several kilometers have been observed to dissipate because of convective circulation eddies developing in clear areas on the edges of the fog and the process spreading through the fog layer (Gurka, 1978).

4.6.4 Effects of Air Pollution

The concentration of aerosol or a polluted atmosphere can also have an effect on fog dissipation. Forkel et al. (1987) indicate that fog dissipation occurs more quickly in a polluted atmosphere because of the enhanced solar heating rates (more absorption) caused by the larger concentration of aerosol.

The effects of different physico-chemical properties of urban, rural, and maritime aerosols on fog dissipation have been reproduced by a one-dimensional radiation fog model of Bott (Bott et al., 1990; Bott, 1991). The total number concentrations for the three cases is 53,880 (urban), 3,890 (rural), and 105 (ocean) particles/cm^3. The total aerosol masses is 145 (urban), 37 (rural), and 28 (ocean) $\mu g/cm^3$. Numerical sensitivity studies using the three different aerosols show that the fog events are completely different, especially for LWC, supersaturation, and fog dissipation. In the urban case, the morning fog was not dissipated by the incoming solar radiation because of the strong radiative absorption of the aerosols. The total solar irradiation reaching the ground in the urban case was 58 W/m^2, while in the rural and maritime case this value was 125 and 183 W/m^2, respectively. These results appear to conflict with the earlier results of Forkel et al. (1987), but may be related to the CCN distribution used in the models.

4.6.5 Windspeed Increase

Saunders' (1973) field studies indicate that in the clearance of radiation fog through insolation and turbulence there is frequently a lifted-fog phase before final dissipation. The likelihood of a lifted-fog phase increases as the geostrophic windspeed increases, especially above 5 m/s. The geostrophic windspeed gives no guidance on the likely cloud base when there is a lifted-fog phase.

5. THE EVOLUTION OF ADVECTION FOG

The evolution of advection fog has not been differentiated into specific evolution stages or phases as has been done with radiation fog. Advection fogs may develop by different advection processes, but the classical process is with the advection of warm, moist air over a colder surface, whether it be land, water, snow, or ice. This type of advection fog may be common off the shores of Novia Scotia (Fitzgerald, 1978), Scotland (Findlater et al., 1989), California (Mack et al., 1974), and other locations (see table 3). Pilie et al. (1979) defined different types of marine fogs along the California coast as (1) fog triggered by instability and mixing over warm water patches, (2) fog developed as a result of lowering (thickening) stratus clouds because of radiational cooling, (3) fog associated with low-level mesoscale convergence, and (4) coastal radiation fog advected to sea via nocturnal land breezes. Of interest, is the statement by Mack and Rogers (1976) that fogs formed by direct cooling from below have not been observed off the west coast.

The evolution of advection fog has not been divided into different stages in the literature; however, this report will arbitrarily and for convenience divide the evolution of advection fog into four stages: (1) precursor, (2) initiation, (3) mature, and (4) dissipation. Results from various fog models and theoretical

and field studies will provide information about these stages and also on other aspects of advection fog that may not fit within these arbitrary categories. Figure 2 is a schematic illustrating the different physical mechanisms and processes involved in an advective fog formed by warm, moist air moving over a colder surface.

5.1 Precursor Stage

5.1.1 Trajectory of the Air Mass

In a study of west coast advection fogs, Goodman (1977) showed that the microphysical structure of the fog is largely determined by the influence of synoptic-scale features. The differences appear to be closely related to the trajectory and air mass history before arrival at the fog site.

Maritime (westerly) trajectories generally resulted in low droplet concentrations (average 89 cm^{-3}), large mean diameters, and broad drop-size distributions. However, maritime trajectories that are modified by continental influences have higher droplet concentrations (average 265 cm^{-3}), smaller mean diameters, and narrow drop-size distributions with sharper peaks.

According to Hudson (1980) and Twomey and Wojciechowski (1969), CCN are sparse in maritime air (~ 100 cm^{-3}), more plentiful in continental air (~ 1,000 cm^{-3}), and abundant in urban air (~ 5,000 cm^{-3}). These CCN serve as particles for condensation of water vapor, and, if numerous, can prevent large supersaturations from being achieved.

5.1.2 Chemical Nature of CCN

In maritime air, the CCN are predominately salt aerosols with hygroscopic properties. The sizes are generally greater than 0.1 μm in diameter and can be as large as 10 μm for the largest salt and dust particles. However, in industrialized areas the chemical composition of atmospheric aerosols is formed from the pollutants of SO_x and NO_x through the gas-phase photooxidation reactions (Hung and Liaw, 1978; 1980). The influence of these type of aerosols has been found to be very important for fog formation and dissipation and the variability in the microstructure of the fog.

A numerical model of Hung and Liaw (1980) was used to simulate how the concentration, particle size, mass of nuclei, and chemical composition would affect the dynamics of advection fog formation. The conclusions of this study are the following:

- For a condensation nucleus to grow into a droplet, the air must have attained a certain degree of supersaturation. However, condensation nuclei associated with a polluted atmosphere can grow into a droplet and produce dense fog without having the air attain supersaturation.

- The major contribution of combustion-related pollutants as condensation nuclei comes from aerosols with species of SO_x followed by NO_x.

• If the mass concentration is kept constant, aerosol distribution with higher particle concentration rather than size of aerosol nuclei makes a greater contribution to the formation of fog.

• If the mass concentration is kept constant, the relative humidity, at which advection fog with visibility below 1000 m is formed, is lower for aerosols of distribution with the higher particle concentrations or smaller size aerosol nuclei.

• More favorable conditions for fog formation are produced by hygroscoptic chemicals with a higher ratio of the Van't Hoff factor (mole number of solution to molecular weight), a higher number density of aerosol nuclei, heavier mass nuclei aerosol particles, and condensation nuclei with larger radii.

5.1.3 Up-Stream Conditioning

Model studies (Rogers et al., 1975; Mack and Rogers, 1976) show that the initial conditions of wind, potential temperature, and mixing ratio should correspond to an air mass that has been conditioned by a significant trajectory over an ocean surface for advection fog to form. Furthermore, they found that the dew-point depression near the surface has to be around 0.5 °C before fog will form after it has been advected over a temperature change of around 4 to 8 °C. However, observations show that fog formation can occur with surface temperature changes as small as 1 °C. Mesoscale convergence and other factors may explain the differences between models and observations.

5.2 Initiation Stage

5.2.1 Advection, Turbulence, and Mesoscale Convergence

In studies involving a two-dimensional model of advection fog (Rogers et al., 1975), results indicated that the advection fog model predicted that a surface horizontal temperature gradients on the order of 4 °C or larger was required for fog formation. Observations indicated that fog formation occurs with surface horizontal temperature gradients as small as 1 °C., and the vertical development of the fog was significantly greater than indicated by the advection fog model.

The apparent discrepancy between the model and observations was attributed to two possible factors that were not included in the numerical model. First, observations by Mack et al. (1975) indicated that mesoscale convergence can be an important element in the initial development and eventual persistence of advection fog. Second, the transfer of heat and water vapor between ocean and atmosphere in response to changes in the ocean surface temperature may proceed at slightly different rates and in certain circumstances may promote fog. Fukuta and Saxena (1973) suggest that the greater rate of molecular diffusion of water vapor than of heat in the laminar sublayer at the ocean surface aids fog formation over warmer water and inhibits fog formation over colder water.

5.2.2 Supersaturation

Saxena and Fukuta (1982) provide theoretical evidence that the evolution of fog supersaturation is inherently time and space dependent. They consider an idealized marine fog formation to be the result of shallow moist air moving over

a colder ocean surface. As the air mass moves over the colder water, the downward transfer of heat and water vapor due to turbulent mixing raises the relative humidity of the air above the condensation level. Fukuta and Saxena (1973) have shown that a maximum in the nominal saturation ratio (the ratio of the total water vapor mixing ratio to the saturation mixing ratio) takes place close to the cold surface and moves away from it with the passage of time because of molecular or turbulent diffusion.

The cold ocean surface contributes to the formation of fog in the warm moist air, and the nominal supersaturation reaches a maximum in the middle of the fog layer; however, the true supersaturation level is much lower since the fog droplet growth reduces the availability of water vapor. The maximum supersaturation occurs close to the upwind fog boundary, but not near the ocean surface because of wind shear and the eventual drying effect (evaporation) of the cold ocean surface.

The possibility of several evaporation zones increases from the effects of nonuniform turbulence and radiative cooling. However, small eddies may also generate local transient supersaturation zones. Thus, zones of droplet evaporation and growth may occur simultaneously in different parts of the fog.

At the instant of fog formation, maximum supersaturation similar to those found in natural clouds may be encountered. As the fog life cycle advances, the turbulent mixing and vapor depletion by growing droplets and by the ocean surface reduces the fog supersaturation.

The maximum supersaturation in the fog is determined by the interaction between the developing nominal supersaturation due to thermodynamic processes and the growth of droplets nucleated in the environment. Therefore, it depends on the droplet growth theory employed. Two models can be applied--a Maxwellian and a non-Maxwellian. According to Saxena and Fukuta (1982), the Maxwellian model of droplet growth is

$$r_i(dr_i/dt) = G(S - ar_i^{-1} + br_i^{-3}), \tag{3}$$

where i is the class containing N_i droplets of radius r_i, G describes the dependence of the droplet growth rate on environmental temperature and pressure, and a and b describe the growth rate dependence on droplet surface curvature (the Kelvin effect) and solute concentration.

The rate at which the actual saturation ratio, S, will change in the fog is dependent upon the rate of change in S_n (nominal saturation ratio) and the growth of the fog droplets.

$$dS/dt = dS_n/dt - A\Sigma N_i r_i^2 (dr_i/dt), \tag{4}$$

where A is the constant of proportionality, including vapor depletion and heating effects.

The non-Maxwellian model contains a radius-dependent term in G of equation (3). Equations (3) and (4) were evaluated numerically for both Maxwellian and non-Maxwellian models for computed supersaturation. The non-Maxwellian theory (more realistic) predicts a retardation of droplet growth and a greater maximum supersaturation. Thus, if the fog droplets grow according to this model, allowing larger maximum supersaturations to be realized, the fog will consist of more but smaller droplets, resulting in a lower visibility (denser fog).

The fluctuating supersaturation (S) in fogs makes it unrealistic to assign or deduce a constant value to S in fogs as done by several fog models and fog prediction schemes such as Fitzgerald (1978), Rogers et al. (1975), Hudson (1980), Meyer et al. (1980), and Fitzjarald and Lala (1989). A constant value of S permits a one-to-one correspondence to be made between the CCN spectra and activation and growth of drops. However, a fluctuating value of S will activate CCN over a broad range in the activity spectrum. Gerber (1991) proposes that the nongradient mixing mechanism (see section 4.4) can be an effective means of activating new CCN entrained into fogs, as well as CCN already located in the interior of fogs.

5.2.3 Turbulent Mixing

Modeling work has shown that fog formation is possible over the ocean by cooling in a shallow layer near the ocean surface. Rogers et al. (1975) indicate that in some locations, for example, the ocean west of California, the large horizontal temperature discontinuities in surface water temperature are not always observed. They and Mack et al. (1973) report on a different process where local fogs are observed to form in cool, nearly saturated air advecting over warmer water. The turbulent exchange and enhanced evaporation from the sea surface result from mixing of warm, moist surface air with cool, moist air at higher levels and from initial condensation in a shallow layer. Radiative cooling of this thin layer lifts the inversion base from the sea surface, and further fog development is promoted by radiative cooling and enhanced mixing beneath the locally induced, low-level inversion.

5.3 Mature Stage

As an advection fog reaches maturity, additional physical mechanisms such as droplet sedimentation, coalescence, scavenging, and evaporation will play a role in determining whether the fog will reach an equilibrium status, vacillate between development and dissipation, or dissipate entirely.

5.3.1 Coalescence and Scavenging

Mack et al. (1973) and Rogers et al. (1975) report that coalescence and scavenging may be important factors in mature advection fogs. Their observations indicate that a continuous light drizzle (drops up to 100 μm radius) frequently accompanies the advection fogs at Vandenberg, California. The presence of large drizzle drops was due in part to the presence of large sea-salt nuclei and also to the fact that the Vandenberg advection fog is generally a low-lying, aged stratus of considerable depth (> 200 m), thus increasing the likelihood of droplet coalescence. Drizzle has been shown to play an important role in the downward transport of moisture in the evolution of fog from lowering stratus and deep advection fogs (Mack et al., 1974).

Fog drops can also scavenge particles since only about 10 percent of the particles act as CCN. The remaining particles can be scavenged by droplets. The most significant effect of the scavenging process is in the resulting alteration of the CCN spectrum after evaporation of the fog droplets. The possible effects may be:

- A subsequent fog formed on the altered CCN spectrum may possess significantly different microphysical features.

- Previously active particles may possibly enlarge, and the CCN spectrum may shift to lower critical supersaturation.

- Scavenged surface active substances may have a contrary effect by retarding the nucleation of fog droplets.

- Alteration of the CCN spectrum may cause droplets to evaporate completely and the residue to form several smaller particles.

5.3.2 Fog Droplet Sedimentation

Model studies from the CALSPAN group (Rogers et al., 1975) evaluated the effects of drop sedimentation and turbulent intensity within an advection fog. Comparison of two steady-state fog models indicated that while larger turbulent intensity may favor fog growth a subsequent larger drop sedimentation may inhibit the growth process, and the net effect may be only a slightly higher fog. They concluded that drop sedimentation has a profound effect on fog growth, and the fall velocity of fog droplets must be modeled carefully in numerical simulation.

5.3.3 Turbulent Mixing and Radiation Cooling/Heating

In the case of a warm, moist air mass being transported over a relatively cold sea initially, once the fog is formed it may reach an equilibrium state. This equilibrium condition is the result of radiation cooling and heating, turbulence, evaporation, and precipitation.

According to Findlater et al. (1989), long-wave radiation from large-scale advection fog gradually takes over the cooling of the air mass, and eventually depresses the fog temperature below the sea surface temperature. This temperature difference initiates convective and radiative heat input from the sea surface, and entrainment of warm air at fog top eventually balances the radiative loss. The radiative loss also leads to increasing LWC due to condensation and increasing evaporation from the sea surface until balanced by increasing precipitation (drizzle) reaching the surface. This equilibrium state is then diurnally modulated by the direct absorption of solar radiation within the fog.

The time scale of adjustment to equilibrium appears to be a few hours and increases with fog depth. Once established, it appears that this equilibrium state could persist almost indefinitely in the absence of major synoptic-scale disturbances.

Another mechanism for promoting fog development and persistence is described in the field and modeling studies of the CALSPAN group (Mack et al., 1975; Rogers et al., 1975). If cool, nearly saturated air is advected over large patches of

warmer water, the subsequent turbulent exchange and enhanced evaporation from the sea surface lead to mixing of warm, moist surface air with cool, moist air at higher levels and condensation. Radiative cooling of the layer lifts the inversion, and further fog development is promoted by radiative cooling and enhanced mixing beneath the inversion.

5.4 Dissipation Stage

5.4.1 Surface Heating

One mechanism for dissipating advection fogs near coastlines is the advection over warmer surfaces heated by solar radiation or heated artificially. Model studies by Mack et al. (1973) indicate that a 200-m deep fog, with LWC in the range 0.25 to 0.3 g/m^3 when advected over a 5 °C temperature rise in the surface, will completely dissipate in slightly over 8 km. During the dissipation, the fog top moves upward apparently in response to the increase turbulence that accompanies the development of an unstable temperature stratification over the warmer surface (thermal internal boundary layer). However, they found that fog movement over a warmer surface is not necessarily sufficient for complete dissipation unless the heat from the surface is distributed throughout the fog layer.

The diurnal nature of coastal advection fog has been documented by several observations and studies such as the recent studies at Vandenberg Air Force Base by the Naval Postgraduate School (Skupniewicz et al., 1991).

5.4.2 Synoptic Disturbance

The persistence of coastal advection fogs is well-known in many parts of the world where up-welling ocean conditions occur (Cereceda, 1991). Findlater et al. (1989) indicate that an advection fog in an equilibrium stage could possibly persist indefinitely if it were not for synoptic disturbances. A storm or frontal passage changes the wind, cloudiness, radiation, temperature, and other variables to the extent that conditions are not favorable for fog.

6. TERRAIN EFFECTS

Valley radiation fogs are a common occurrence in many places of the world. It is not surprising that some of the field experiments have addressed radiation fogs in valleys.

6.1 Valley Fog Formation

Pilie et al. (1975a, b) explain the formation of radiation fog at Elmira, New York (Chemung River Valley), by contributions from nocturnal valley circulations described by Defant (1951) and many others. The nocturnal valley circulation consists of downslope winds, an elevated upvalley return flow near the center of the valley, and a lower downvalley wind. Pilie et al. suggest that the downslope winds and the upvalley return flow near the center of the valley assist in the development of an upper-level temperature inversion and also help in conditioning the low- to mid-levels of the valley by radiative cooling. During this period, dew deposition at the cold surface creates a low-level dew-point inversion. The downvalley wind that forms later provides continuity for the downslope wind,

while restricting the upward motion of air near the valley center. Temperature and dew-point converge at mid-levels as the dew-point inversion deepens and a thin layer of fog forms aloft. Once fog has formed aloft, the fog is propagated downward because of radiative flux divergence at the fog top, the increased mixing of cold foggy air with clear, almost saturated air beneath the fog top, and the increase in instability in the lower levels.

Pilie et al. report that on several occasions, fog formed following sunrise. Surface warming and dew evaporation caused vertical mixing of moist low-level air with cooler air in the valley center aloft. Adiabatic cooling of rising air parcels contributes to the fog formation.

Fitzjarrald and Lala's (1989) analyses of various data in the Hudson River valley lead them to postulate that the nocturnal boundary layer consists of three important dynamic layers: (1) a stable surface layer extending to 20 m in which turbulent activity is intermittent, (2) a boundary layer in which channeling can occur if the pressure gradient along the valley is large enough and the depths reaches 150 m and (3) blends into a region in which the wind direction is close to that of the synoptic-scale geostrophic wind. If the channeling in the boundary layer consists of moist southerly flow, dry westerly advection from deterring fog onset can be prevented, at least in the Hudson River valley.

6.2 Valley Fog Dissipation

Pilie et al. (1975) described the valley fog dissipation process as follows: a subsidence develops in the valley center in response to solar radiation heating the valley slopes causing slope winds. Any elevated hills in the valley are the first regions where the fog dissipates because the optical depth of fog is least over the hills. Solar insolation penetrates the thinning fog layer, thus warming the higher ground. As the hills become exposed, slope circulations develop, with corresponding compensating sinking motion over the neighboring foggy lower terrain. The resultant adiabatic warming evaporates the fog and dissipation proceeds rapidly.

Plank and Spatola (1976) describe the natural dissipation of fogs in two Applachian valleys by photography. Other measurements were limited. The first valley, Howard Creek Valley, is a narrow (width from ridge to ridge ~ 5 km) and steep-walled valley with numerous intersecting canyons and ravines. The fog formed after midnight and was around 244 m deep. Initially, the fog top had a flat, stable appearance with only minor undulations, but the fog structure changed around 0830 LST. At this time, the fog along the edges of the valley moved slowly upward into the various canyons and ravines that intersected the valley sides. The upper surface of the fog along the valley centerline also seemed to be gradually subsiding. By 0845 LST, the upper surface of the fog contained within the peripheral canyon and ravines and that adjacent to the valley walls, began to have a cumuliform, cellular appearance. This cumuliform structure propagated toward the valley centerline and extended completely across the valley by 0900 LST. The surface visibility within the fog had increased to over 1 km by this time. The first clearing occurred between the cumuliform elements along the valley walls, and this type of clearing progressed from the walls toward the valley centerline. The cumuliform elements then began to evaporate and disappear, creating larger and larger cleared areas between remaining elements. Finally, the residual elements dissipated, and the valley became completely clear by 0935 LST.

The second valley, Greenbrier Valley, is a wider valley (~ 13 km) with a generally smooth valley floor. The fog formed around 0330 LST and was initially about 61 m deep, but increased to around 107 m by 0645 LST (such an increase in the fog depth near sunrise was a commonly observed phenomenon during this project period). By 0918 LST the fog top had a well-defined convective structure that was composed of rows of cumuliform elements that were oriented in a general north-south direction and were spaced about 700 m apart. This cumuliform structure was not apparent at the surface level. Cleared strips developed between the rows of convective elements after 0925 LST but did not extend to the surface level. However, the fog at the surface began to lift above the ground at this time period. The strips widened and cleared spaces appeared between the elements along the rows. Eventually, the cloud elements began to evaporate and disappear until it became completely clear around 0954 LST. The bases of the cumuliform elements lifted continuously in altitude during the dissipative phase, but the average top level of the clouds remained about the same.

The difference in the fog dissipation between these two valleys is described in terms of topography differences. In the narrow valley, an apparent daytime valley breeze and upslope winds were contributing factors to the dissipation of the fog. While the fog of the wide valley dissipated convectively, the valley breeze and upslope winds were lacking, and it was suggested that the convective nature of the fog in the wider valley was correlated with the topography within the valley. However, an examination of wind directions and the wavelength-to-height ratio (~ 3:1) of the cloud elements suggests that roll convection may have been responsible, in part, for the fog dissipation.

Other studies of valley fogs have been reported by Bonner and White (1972) and Mack et al. (1973), but these studies were conducted at coastal sites, and the fog developments were combinations of advective, orographic, and radiative fogs.

6.3 River and Lake Fogs

The effect of large rivers on the formation of radiation fogs has been investigated in China by Qian and Lei (1988, 1990). They used a two-dimensional numerical model to study the relationship of fog over the Changjiang and Yangtze Rivers.

The results from their modeling efforts showed the following relationships:

- The wind field is coupled to the terrain or river banks. Steep river banks (> 0.08) are not conducive for radiation fog because of the unfavorable wind fields developing along the slopes and valley.

- The temperature differences between the river and land governed the strength of circulation over the river. A small temperature difference resulted in a smaller and weaker circulation.

- The width of the river governed the availability of moisture. A wider river provides a larger moisture source and results in a deeper fog.

• Fog over or near rivers can reach a high level (~ 500 m) if the river is wide (~ 1000 m) and the slopes along the banks are not too steep (< 0.08).

Advection-radiation fogs were investigated near Lake Michigan by Ryznar (1977). The occurrence of fog is more frequent during the summer months, and the frequency of occurrence decreases with distance from the lake. Locations near the lake shore have about twice as many hours of fog as locations some 19 km away. A combination of onshore gradient winds and lake breezes acts to maintain higher average relative humidities near the shoreline. Terrain differences are of secondary importance.

A preliminary study of a cold advection warm fog over a small reservoir in Tennessee was conducted by Connell (1979) during November 1977. The results were limited to one case study. The onset and maintenance of the lake fog by advection of cold air may have been related to a small acoustic echo region that descended to lower heights above the main inversion and the decreasing turbulence intensity. Other observations included noting a low and simple inversion structure over the lake and weak thermal activity in the boundary layer. Also, the horizontal gradients of lake temperature or shore breezes, or both, can strongly affect fog over the lake.

7. SUMMARY AND CONCLUSIONS

A review of the results of radiation and advection field programs and numerical models from the scientific literature has revealed several areas of further study. These areas will be briefly discussed below.

7.1 Radiation Fog

In the precursor stage of radiation fog formation, the synoptic and boundary-layer conditions preceding fog formation are important to investigate and understand since knowing what conditions are favorable for fog formation will allow establishment of better initial conditions for fog models and also allow prediction of fog or no fog situations. In the sunset and conditioning stage, turbulence, CCN, radiative cooling, and the formation of dew are critical factors in determining fog onset time. There still seems to be uncertainty whether radiation fog begins at the surface, aloft, or both. A number of model studies appear to favor fog aloft formation, but this may be a reflection of the grid resolution of models. Field studies report both methods of formation. Air pollution affects the size and number distribution of CCN that can determine the degree of supersaturation. Some progress has been made on these aspects.

As the fog enters the mature stage, questions arise concerning fog droplet spectra, supersaturation, fog droplet settling, deposition, and scavenging. Advection, evaporation, and variability of supersaturation affect the spatial and temporal variation of the fog and also its equilibrium state. These effects have not been studied very effectively because of the limited dimensions of most models.

At the sunrise period, questions arise concerning the necessary conditions for early morning fog formation or initial dissipation. The question here is why does fog form at sunrise at times and at other times start to dissipate.

A number of physical mechanisms have been identified that lead to fog dissipation such as solar radiation heating, advection of cloud cover, convective mixing, and increasing windspeeds. However, it has been reported that the mode of clearance may at times be from the ground upward and sometimes propagates downward. The factors explaining the difference in mode of dissipation need further study. The role of air pollution aerosols on the dissipation of radiation fog has been studied by Forkel et al. (1987) and Bott (1991). However, their results appear to conflict with each other, but may be related to the CCN distribution used in their models.

7.2 Advection Fog

The field studies of Pilie et al. (1979) indicated some different mechanisms that could lead to the formation of advection fog such as mixing over warmer water, thickening of preexisting elevated stratus, mesoscale convergence, and advection of coastal radiation fog. However, most studies have concentrated on the basic mechanism of advection of warm, moist air over a colder surface.

The precursor stage of this type of advection fog depends strongly on the history of the air mass and its up-stream conditioning. As it proceeds to the initiation stage, the role of turbulence, radiation cooling, supersaturation, and evaporation determines the spatial and temporal variability of the fog formation. The differences between the transfer of heat and water vapor are also important and may need further investigation in the laboratory and in the field, and possibly by models. In the mature stage, the coalescence, scavenging, and droplet sedimentation processes appear important factors in determining whether an advection fog can reach an equilibrium state.

The dissipation of advection fog as it moves inland is well-known, but there appears to be a lack of in-depth studies concerning the evolution of advection fogs over land surfaces and their dissipation. However, this may be reflected, in part, by the limited nature of the survey.

A limited number of studies on radiative-advection fogs in valleys and over rivers and lakes were found and reviewed. Terrain can exert a strong effect on fog formation, persistence, and dissipation. Further studies are needed on these aspects.

One additional factor that is difficult to assign to a particular stage of fog evolution should be mentioned. A study by Faulkner (1978) implies that large-scale climatic changes may be related to long-term incidence of radiation and advection fogs. A decline in radiation/advection fog at Vancouver, Canada, between 1940-1970 was significantly correlated with changes in the large-scale atmospheric 500-mbar circulation patterns during the same period.

In conclusion, field studies and numerical models have progressed considerably in understanding and modeling radiation and advection fog formation and dissipation. The evolution of these fogs has been generally described by different stages or phases. Several of these aspects need further study, especially in order to understand the intermittency and spatial and temporal variability of these types of fogs and eventually their dissipation.

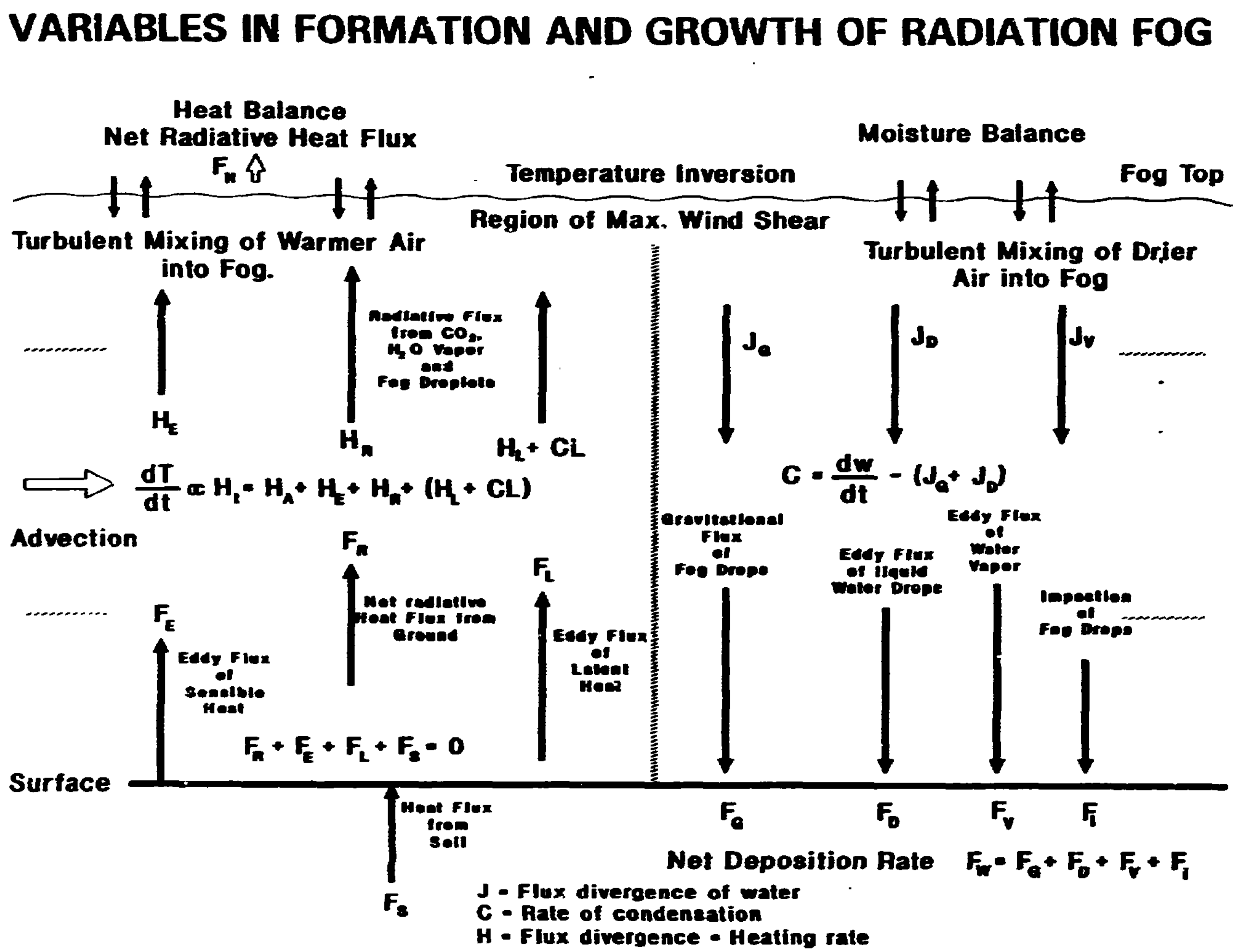
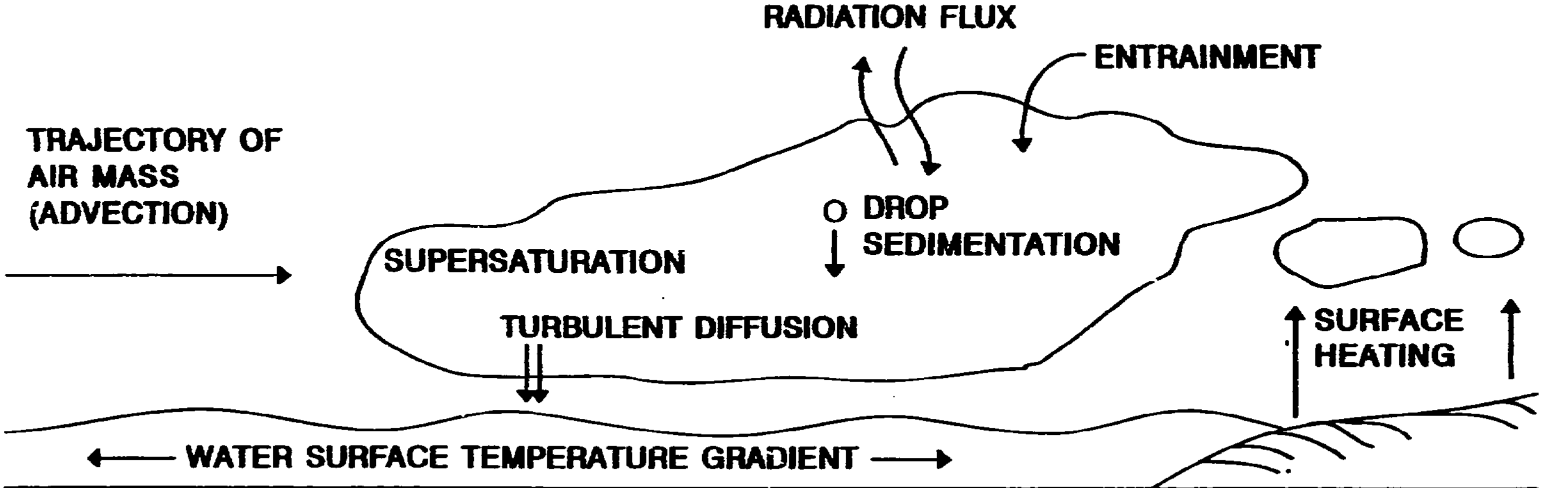


Figure 1. Mechanisms and processes involved in radiation for evolution (adapted from Mason, 1982).



• Trajectory of air mass • Chemical nature of CCN • Dewpoint depression (<0.5 °C)	• Advection • Mesoscale convergence • Temperature gradient 1-8 °C • Turbulent diffusion (heat & water vapor) • Supersaturation	• Radiation flux • Entrainment (evaporation) • Coalescence and scavenging • Areas of warmer water (turbulent heat & moisture transfer) • Increasing fog depth	• Synoptic disturbance • Surface heating • Evaporation
PRECURSOR STAGE	INITIATION STAGE	MATURE STAGE (Equilibrium State)	DISSIPATION STAGE

Figure 2. Mechanisms and processes invloved in advection fog evolution.

TABLE 1. RADIATION FOG FIELD PROGRAMS FROM 1957 TO PRESENT

Investigators	Yr of Pub	Location
Duynkerke	1991	Cabauw, Netherlands
Fitzjarrald and Lala	1989	Hudson River Valley, NY
Guedalia et al.	1988	Valladolid, Spain
Fuzzi et al.	1983-1988	Po Valley, Italy
Meyer et al.	1983-1986	Albany, NY
Findlater	1985-1987	Cardington, England
Wessels	1984	Cabauw, Netherlands
Choularton et al.	1981	Meppen, W. Germany
Meyer et al.	1980	Albany, NY
Caughey et al.	1978	Cardington, England
Plank and Spatola	1976	Appalachian Valleys
Roach et al.	1976	Cardington, England
Pilie' et al.	1975a, b	Chemung Valley, NY
Chisholm and Kruse	1974	Hanscom AFB, MA
Mack and Pilie'	1973	Travis AFB, CA
Mack et al.	1973	Vandenberg/L.A., CA
Garland et al.	1973	Cardington, England
Pilie et al.	1972	Elmira, NY
Bonner and White	1972	Redwood Valley, CA
Davis	1957	Seabrook Farms, NJ

TABLE 2. REVIEW OF VARIABLES MEASURED DURING SOME RADIATION FOG FIELD PROGRAMS

Variables	B-W 1972	PEMRK 1972	MERKP 1973	M-P 1973	PMKER 1975	RBCGR 1976	P-S 1976	CDC 1978	CFLMSS 1981	W 1984	F 1985	MLJ 1983-6	F-L 1989
Aer.Part/Cond.Nuclei	X	X	X	X	X	X						X	
Super Saturation												X	
Drop-Size Distribution	X	X	X	X	X	X			X			X	
Droplet Growth									X				
Liquid Water Content		X	X	X	X				X			X	X
Solar Radiation							X						
Radiation Fluxes		X		X								X	X
Visibility	X	X	X	X	X	X	X					X	X
Droplet Settling									X				X
Turbulence/Entrain.													X
Diffusion(Turbulent)						X			X				X
Heat Budget						X			X			X	X
Moisture Budget						X			X			X	X
Dew Formation		X	X	X	X							X	X
Evaporation					X							X	
Vegetation(Impaction)												X	
Cloud Cover									X				
Synoptic Coupling													X
Advection			X	X		X							X
Cellular Elements									X				
Quasi-Periodic Oscil.						X			X				
Terrain Effects	X		X		X		X						X
Precipitation													
Temperature	X	X	X	X	X		X					X	X
Wind Measurements		X	X	X	X		X					X	X
Dew Point(R.H.)	X	X	X	X	X								X
Vertical Motion		X	X	X	X								X
Photography							X						

TABLE 3. ADVECTION FOG FIELD PROGRAMS FROM 1973 TO PRESENT

Investigators	Yr of Pub	Location
Medvedev and Nikel'shparg	1991	Odessa, Ukraine
Tsujinaka and Nakata	1986	Sakata Bay, Japan
Woodcock	1984	Cape Cod Canal, MA
Gavrish	1981	Latvian Coast, Latvia
Woodcock et al.	1981	New England Coasts
Dushkin	1981	Minsk, Russia
Dushkin	1980	Moscow, Russia
Connell	1979	Woods Resvior., TN
Kunkel	1980-1984	AFGL WTF, MA
Pilie' et al.	1979	California Coast
Hoppel and Fitzgerald	1977	New Foundland Coast
Ryznar	1977	Lake Michigan, MI
Goodman	1977	San Francisco Pen., CA
Mack and Katz	1976	Nova Scotia Coast
Rogers et al.	1975	Project Fog Drops, CA
McClure	1974	California Coast
Mack et al.	1974	Project Sea Fog, CA/NS
Mack et al.	1973	L.A./Vandenberg, CA

TABLE 4. LISTING OF RADIATION FOG MODELS AND THEIR AUTHORS

1) Zdunkowski and Associates (1991-1966)

   a) Bott (1991)
   b) Bott et al. (1990)
   c) Forkel et al. (1990)
   d) Bott et al. (1988)
   e) Forkel and Seidel (1988, 1989)
   f) Forkel (1987)
   g) Welch et al. (1986)
   h) Forkel et al. (1984)
   i) Forkel and Zdunkowski (1983)
   j) Zdunkowski et al. (1982)
   k) Zdunkowski et al. (1980)
   l) Zdunkowski and Welch (1976)
   m) Zdunkowski and Barr (1972)
   n) Korb and Zdunkowski (1970)
   o) Zdunkowski and Nielson (1969)
   p) Zdunkowski et al. (1969)
   q) Zdunkowski et al. (1967)
   r) Zdunkowski et al. (1966)

2) Duynkerke (1991)

3) Qian and Lei (1990)

4) Musson-Genon (1989)

5) Vehil (1989)

6) Pandis and Seinfeld (1989)

7) Vehil and Bonnel (1988)

8) Wobrock (1988)

9) Smolarkiewicz and Fitzjarrald (1988)

10) Qian and Lei (1988)

11) Brown, Roach and Associates (1987-1976)

   a) Turton and Brown (1987)
   b) Mason (1982)
   c) Brown (1980)
   d) Brown (1979)
   e) Roach and Slingo (1979)
   f) Brown and Roach (1976)
   g) Roach (1976)

TABLE 4. (cont.)

12) Musson-Genon (1987)

13) Zhang et al. (1987)

14) Zhou (1987)

15) Ohta and Tanaka (1986)

16) Bykova (1986)

17) Hacker-Thomas (1985)

18) Latham (1981)

19) Oliver et al. (1978)

20) Buykov and Khvorost'yanov (1977)

21) Zakharova and Sedunov (1976)

22) CALSPAN Models (1975-1972)

    a) Lala et al. (1975a, b)
    b) Pilie' et al. (1972)

23) Zakharova (1973)

24) Melkaia (1968)

25) Fisher and Caplan (1963)

TABLE 5. RADIATION FOG MODELS AND VARIABLES CONSIDERED IN DEVELOPMENT

Variables	F-C 1963	ZNK 1967	Z-N 1969	K-Z 1970	PEHRK 1972	Z-B 1972	MERKP 1973	LNJ 1974	Z-V 1976	B-R 1976	OLW 1978	R-S 1979	B 1980	M 1982	ZPMK 1982	FPMZ 1984	WRC 1986	LNG 1987	T-B 1987	S-F 1988	LNG 1989
Dimension (S)	1/2	1	1	1	1	1	2	1	1	1	1/2	1	1	1	1	1	2	1	1	1/2/3	1
Gas/Vapor Phase		X	X	X		X		X	X	X	X	X	X	X	X	X	X	X	X	X	X
Spher. Aerosol		X		X		X			X	X		X	X	X	X	X	X	X	X		X
Irreg. Aerosol																					
Spctrl Regn/Line		X		X					X			X			X	X	X			X	
Transmission		X		X		X			X	X	X	X	X	X	X	X	X	X	X	X	X
Single Scatterng		X		X					X			X		X	X	X	X	X			X
Multi Scattering		X		X					X			X		X	X	X	X	X			X
Thermal Emission		X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X
Absorption		X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X
Extinction/VSBY				X									X	X			X	X	X		X
Drop Size Spectr		X	X	X		X	X		X	X		X	X	X	X	X	X	X	X		X
Droplet Growth													X	X							
Aerosol Sz Spctr													X	X	X	X	X				
Plane Paral Atms		X		X					X						X	X	X	X			X
Phase Function		X		X					X												
Solar Spectrum		X		X					X	X	X		X	X	X	X	X	X			X
Diffuse Skylight		X		X					X						X	X	X				
Srfc Albedo/Emis				X						X		X	X	X	X	X	X		X		
Polarization																					
Liquid Wtr Cntnt	X	X	X	X	X		X		X	X	X	X	X	X	X	X	X	X	X	X	X
Condensation		X							X								X			X	
Supersaturation	X		X		X	X	X			X			X	X	X	X	X	X	X		X
Droplet Settling					X		X		X	X			X	X		X	X	X	X		X
Dew Formation					X			X						X			X				
Evaporation	X		X		X		X		X						X	X	X	X			X
Turbulent/Diffsn	X	X	X		X	X	X	X	X	X	X		X	X		X	X	X	X	X	X
Temperature	X	X	X		X	X	X	X	X	X	X	X		X		X	X	X	X	X	X
Advection	X						X				X			X		X	X	X	X	X	X
Clouds										X	X	X		X	X	X					
Precipitation																					
Soil Heat Trans		X	X		X	X		X	X	X			X	X		X	X	X	X	X	X
Soil Moist.Trans										X									X		
Vegetation																					
CouplingDynamics																					X
CellularElements																				X	
Quasi-Period Osc																	X				
Terrain Effects																					
Validation							X			X	X	X		X	X			X	X		

TABLE 6. LISTING OF ADVECTION FOG MODELS AND THEIR AUTHORS

1) Sun et al. (1991)

2) Ohnogi and Shibata (1986)

3) Khvorost'yanov (1982)

4) Buikov et al. (1981)

5) UAH (University of Alabama in Huntsville) Models (1982-1978)

   a) Hung and Liaw (1982)
   b) Hung and Liaw (1981a, b)
   c) Liaw (1980)
   d) Hung and Liaw (1980)
   e) Hung and Vaughan (1979)
   f) Hung et al. (1979)
   g) Hung and Liaw (1978)

6) Matveev and Soldatenko (1978)

7) Fitzgerald (1978)

8) Rao et al. (1977)

9) Barker Models (1984-1975)

   a) Weyman (1984)
   b) Barker (1977)
   c) Barker (1975)
   d) Barker and Baxter (1975)

10) CALSPAN Models (1976-1973)

   a) Mack and Rogers (1976)
   b) Mack and Katz (1976)
   c) Rogers et al. (1975)
   d) Eadie et al. (1975)
   e) Mack et al. (1973)

11) Pepper and Lee (1973)

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